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Formation and dynamics of magma reservoirs

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Abstract

The emerging concept of a magma reservoir is one in which regions containing melt extend from the source of magma generation to the surface. The reservoir may contain regions of very low fraction intergranular melt, partially molten rock (mush) and melt lenses (or magma chambers) containing high melt fraction eruptible magma, as well as pockets of exsolved magmatic fluids. The various parts of the system may be separated by sub-solidus rock or be connected and continuous. Magma reservoirs and their wall rocks span a vast array of rheological properties, covering as much as 25 orders of magnitude from high viscosity, sub-solidus crustal rocks to magmatic fluids. Timescales of processes within magma reservoirs range from very slow melt and fluid segregation within mush and magma chambers and deformation of surrounding host rocks to very rapid development of magma and fluid instability, transport and eruption. Developing a comprehensive model of these systems is a grand challenge that will require close collaboration between modellers, geophysicists, geochemists, geologists, volcanologists and petrologists.

Introduction

The modern concept of a magma chamber can be traced back more than 100 years. Daly (1) summarised several decades of developing ideas about volcanism, its connection with the deep sources of magma in the Earth's Interior, the relationship to plutonic rocks and evidence for the existence of large subterranean bodies of magma from geological evidence, notably the formation of calderas. The ideas of forceful intrusion of magma, magma differentiation by separation of melt from crystals and development of large bodies of mostly molten rock progressively gained currency. By the time G.W. Tyrrell wrote his classic textbook on petrology (2) in 1926 almost all the modern notions of magmatic systems had been introduced at least in outline, including the idea of extracting melt from what was then termed 'crystalline mesh'. The concept of the magma chamber supplying volcanoes and within which igneous differentiation happens became the main paradigm for volcanology and igneous petrology. Complementary lines of enquiry developed in parallel with the magma chamber concept, one being the geological and geophysical evidence for formation of magma chambers by intrusion and the other was the recognition that space needed to be created for intrusions to enable voluminous magma chambers to form. To a large extent intrusion and magma chamber formation became synonymous and a large literature developed that identified a variety of intrusion mechanisms (e.g. 3).

Increasing understanding of phase equilibria and igneous differentiation by crystallization was pioneered by Bowen (4), and extended by the discovery of layered intrusions with documentation of their geological and petrological characteristics (e.g. 5-6). These studies established that much of the geochemical diversity of magmas, and the refractory characteristics of many plutonic rocks, must be caused by separation of crystals and melt, albeit complicated in some instances by magma mixing and assimilation of host rocks. The tendency was to assume that this separation took place in magma chambers, such that chemical differentiation and magma chambers became inextricably linked. However, there was also recognition that melts could escape from crystals by diverse processes. One important example is the discovery of adcumulate (refractory) rocks in layered intrusions (7), testifying to efficient separation of melt and crystals, although the mechanisms involved

remain elusive and controversial (8). Two possibilities for adcumulate formation are convective and diffusive exchanges between melts in a mush and a magma chamber (7-9). Another is by mush compaction (10,11). Evidence for melt and crystal separation is also apparent in granites, as exemplified by various hypotheses for melt segregation from the partially melted source regions (12, 13), the recognition that many granites have cumulate structures and compositions (14,15), and existence of large volumes of crystal-poor evolved silicic melts (16).

The open-system nature of magma chambers was also recognised with evidence for recharge by new magma from depth, which was used to explain the very common occurrence of cyclic layering in mafic and ultramafic intrusions (e.g. 17,18), magma mixing phenomena (e.g. 19-21) and triggering of volcanic eruptions (23). In the 1980's much research was devoted to understanding the internal processes within magma chambers once formed (e.g.24,25). The quintessential shallow magma chamber beneath a volcano is depicted in Figure 1.

The magma chamber remains a valid and enduring concept that has been very successful at explaining many igneous and volcanic observations. Indeed, many models of the chemical evolution of igneous rocks are predicated on the magma chamber paradigm, including mathematical treatments of fractional crystallisation (with or without associated crustal assimilation), and so-called liquid lines of descent followed by discrete batches of parental magma undergoing sequential crystallisation. However, apparent inconsistencies have emerged with a number of observations failing to reconcile with the magma chamber concept.

Geophysical studies have struggled to detect the postulated large magma chambers, the observations being much better explained by large volumes of hot igneous rock containing quite small amounts of melt in the crust beneath many active volcanoes and ocean ridges (e.g. 26-28). Caveats concern the resolution of geophysical images ($>10^2$ to 10^3 m) that cannot define the detailed geometry or scale of melt distributions, which strongly affect estimates of melt fractions. Development of tools for precise absolute dating of crystals, U series studies of isotopic disequilibria and estimates of residence times for zoned crystals in magma using diffusion chronometry showed that magmatic systems could have lifetimes of hundreds of thousands to millions of years, but that commonly phenocrysts in volcanic rocks only became suspended in their host magma for periods of years to centuries prior to eruption (29-31). The latter observations indicated that magma chambers are often ephemeral. Conversely, a large spread in radiometric ages of suspended crystals (notably zircon) required very long residence times for crystals in sub-volcanic magma reservoirs. These observations have been reconciled by proposing that magmatic systems are dominated by igneous mushes wherein melt is distributed within a framework of crystals.

Many researchers have thus concluded that magmatic systems are volumetrically dominated by the mushy state (26, 32-34), that voluminous magma chambers are subordinate to mush and that magma chambers, once formed, may be short-lived. Moreover current thinking has shifted towards the idea that magmatic systems develop throughout the crust with differentiation mostly taking place in the lower and middle crust, while the upper crust is largely a region of magma accumulation and consolidation (35).

These notions are reviewed by Cashman et al (36). Figure 1 in Edmonds (this volume) and Figure 2 shows schematically the emerging view of mature magmatic systems in which multiple and ephemeral magma chambers develop within a volumetrically dominant mush system.

In this paper we summarise contemporary understanding of magmatic systems framed in terms of physical properties and dynamics. We outline an emerging paradigm and discuss key scientific challenges for future research.

Some working definitions

Terminology provides an essential pre-requisite for scientific discourse, but can present challenges, for example, when the terminology has genetic connotations; where meaning is implicit rather than explicit; if terms are ambiguous; or if meanings are not fully agreed within a scientific community or change with time. These difficulties are widespread in the igneous and volcanic literature. Here we present our definitions of key terminology so at least it can be clearly understood what we mean, even if there is not necessarily consensus about these definitions.

At the largest scale *a magmatic system* constitutes four interacting physical domains (Figures 3 and 4). The domains are *magma* (melt \pm crystals and exsolved fluids), *mush* that contains melt and associated fluids distributed within a crystalline framework, *super-solidus rocks*, and sub-solidus *host rocks* that are influenced by transfer of magmatic heat, influx of fluids of magmatic origin and magma-induced stress. The host rock domain includes both older country rock and fully solidified cognate intrusions of the magmatic system. Here we recognise that domains in which melt is distributed within a crystalline framework can have markedly different physical properties and therefore dynamic behaviours (Figure 4). We term rocks in which the partial melt fraction is very low (a few percent or much less) and unconnected as super-solidus rock. We define mush as domains where the melt fraction is interconnected with significantly reduced viscosity and strength compared to sub-solidus and super-solidus rocks. The external environment can influence a magmatic system, for example through tectonic processes and the cooling effects of convecting meteoric water. The system might extend from the source of melt generation in the mantle (or within the deep crust) through the crust and usually to the Earth's surface. Some magma solidifies in the crust to form plutonic rocks; other magma erupts to form volcanoes.

We define *a magma reservoir* as the domains within the magmatic system that contain melt (\pm fluid) and by definition are above the solidus. In some mature systems there may be one continuous, interconnected domain (reservoir) from the mantle into the shallow crust while in other systems there may be multiple domains (reservoirs) separated by zones of completely solid (melt-free) rock (Figure 3). We acknowledge that it is common in the literature to regard the terms magma chamber and magma reservoir as synonymous, but this is increasingly problematic because of the recognition that melt distributed at a small scale within largely crystalline rock may under some circumstances be rapidly extracted (e.g. 37-39). Also, it has become apparent from many lines of evidence that most magmatic systems are dominated by crystalline frameworks in which melt (\pm fluid) is distributed either on a micro-scale (Figure 4), typically that of crystal dimensions (34, 40) or on a mesoscale in the form of veins, lenses and pockets. Textures of plutonic xenoliths in volcanic rocks often

testify to the presence and reactivity of such melts (e.g. 41,42). Of course melt-dominated regions can sometimes contain suspended crystals (\pm fluid) within a reservoir, but these regions may be subordinate in volume and may be ephemeral. Our definition of magma reservoir thus includes both crystal-dominated and melt-dominated domains. There is a useful parallel in this terminology with the way reservoir is used in other geological systems where porous rocks host oil, gas and or water.

In the above definition a magma reservoir is characterised by two fundamentally different physical sub-systems or sub-domains. We define a melt-dominated region as a *magma chamber* that is indeed synonymous with the classic text-book connotation. Crystal phases are suspended and exsolved fluids may be present too. We define a crystal-dominated regime as a *mush*, a term now widely used in the petrological and geophysical literature. The crystalline component forms a framework and melt (\pm fluid) is distributed as a network between the crystals. The amount of melt in a mush can vary widely from vanishingly small to high melt fractions which in some circumstances can exceed 60%. We have been unable to find any precise definition of an igneous mush. However, the term has a long vintage, extending back to the “mesh” of Tyrell (2). In materials sciences mush is used in the same way (e.g. 43), whereas what in earth sciences is called a magma is termed in materials science a slurry or liquid (if no suspended solids are present). We also recognise strong and weak mush, the rationale for which is developed in the next section.

Rheology of magmatic system domains and their constituents

We now consider magmatic systems in terms of rheological domains and dynamic regimes that are governed by rheology (Figure 3). Three main domains have been defined: magmas, mush and host rocks. Each domain can contain several multi-component phases including crystalline solids, melts and fluids, although melt is by definition absent in sub-solidus host rocks. The shear viscosities of these domains and their constituent phases span more than 25 orders of magnitude from the high viscosity of hot subsolidus rocks (10^{18} to 10^{20} Pa s) through silicate melts (10^0 to 10^7 Pa s) to the viscosity of magmatic fluids (10^{-2} to 10^{-5} Pa s). Such a huge viscosity range implies a very large range of time scales as a consequence of various instabilities that can arise, related principally to buoyancy. These time scales range from the exceedingly slow to the very fast, encompassing behaviours that may develop over the lifetimes of igneous systems (10^5 to 10^7 years) to the rapid separation of volatiles during explosive eruptions (tens of seconds).

For magma (Figure 3), rheology is controlled principally by composition, dissolved H₂O content and temperature of the (usually silicate) melt phase and the volumetric proportions and physical properties (size, shape and surface tension in the case of volatiles) of suspended crystals and bubbles (44-48). Melt viscosities typically vary from less than 1 Pa s to 10^6 Pa s, although viscosities up to 10^{13} Pa s are relevant in some situations (e.g. very dry silicic melt). Magmas have Newtonian behaviour at modest to low content of bubbles and crystals, but at high concentrations they can develop complex non-Newtonian rheologies (48, 49). Magmas can erupt (50) and can be transported through, and be emplaced within, colder sub-solidus environments. Magma chambers allow crystals, fluids and melts to separate from one another by processes such as crystal settling, bubble rise and crystallization onto the chamber walls (24,25).

The rheologies of super-solidus rocks and mush (as defined in Figure 3) are governed principally by deformation of the crystalline framework (51-53). There are significant differences, however, between a super-solidus rock lacking melt interconnectivity and mush. Melting of crystalline rock significantly reduces shear viscosity by generating an interconnected melt network along grain boundaries, which allows faster diffusive transport. Shear viscosity reductions of one to three orders of magnitude can occur at the onset of melting and melt contents of a few percent (53,54). Melt interconnectivity depends on melt-crystal wetting angle (10,55). Whereas mafic melts can wet crystal boundaries down to very small melt fractions (10), felsic melts have high wetting angles that require a certain threshold melt fraction for connectivity. The threshold melt fraction is found experimentally to be around 7% in dry partially molten granite (56) and has been estimated from theoretical models to be between 1 and 2% for basalts (57). The transition between super-solidus rock and mush is likely to be affected by strain rate, with connectivity being enhanced at higher strain rates. At the “Melt Connectivity Transition” (MCT) there is a large loss of strength and reduction in effective viscosity by about two orders of magnitude. The MCT is also important for the development of matrix permeability for melt flow (56) and for faster diffusive transport for mush deformation by dissolution-precipitation mechanisms. We suggest that the MCT provides a rationale for dividing the domains with very low melt fractions (super-solidus rock) and those with higher than a (small) threshold melt fraction (mush) respectively (Figure 3).

The rheology of mush is complex and a topic of much contemporary debate, particularly regarding processes in the mush domain that control many aspects of igneous petrogenesis and dynamic behaviours, as discussed in more detail below. At low melt fractions the deformation mechanisms of mush are likely to be similar to hot, solid igneous rocks. Two main mechanistic deformation regimes have been recognised in hot rocks (52,54,58), namely dislocation creep and dissolution-reprecipitation. The former has a highly non-Newtonian power law rheology whereas the latter has a Newtonian rheology. Holness et al (8) have shown that many cumulate plutonic rocks lack features of internal crystal distortion and recrystallization (e.g. sub-grain development), while preserving fabrics acquired during magma flow and crystal settling. Features of dislocation creep are, however, observed in some large intrusions like the Bushveld, South Africa (8). It is common to characterise a non-Newtonian material by an effective viscosity (at small strain rates). Effective viscosity is influenced by melt content. Mush effective shear viscosities are expected to be in the range 10^{18} to 10^{13} Pa s and to decrease with increasing melt content and increasing temperature for both the cases of Newtonian and Non-Newtonian rheologies. Mushes can also deform heterogeneously with localisation of strain along shear zones accompanied by melt segregation (e.g. 59,60). In mush compaction the dominant factor is bulk viscosity rather than shear viscosity (61). Whereas shear viscosity remains relatively constant until high melt fractions are reached, bulk viscosity can decrease by two to three orders of magnitude as melt content changes from very small to high (62). For example a typical bulk viscosity might change from 10^{18} Pa s at very low melt fractions to 10^{14} Pa s at 40% melt fraction (47).

At high melt fractions (Figure 3) mush rheology may be rather different. Loading stresses are transmitted through force chains of crystals in local contact (63). Local stresses at crystal contacts can vary greatly. Force chains may give the mush some yield strength which may

need to be exceeded to initiate shearing (e.g. 64). Once the yield stress is reached via disruption of the force chains, melt-rich mush can rapidly transition to crystal-rich magma (63,65,66).

Mushes themselves cannot erupt, except as entrained xenolith fragments (36) or under special circumstances at very shallow depths. However, disrupted mush can be a significant component of some crystal-rich magmas in the form of antecrysts and glomerocrysts (e.g. 67-69). Mechanisms of melt (and fluid) segregation from crystals within a mush involve porous media and reactive flow, which can also drive chemical differentiation (e.g. 41,70-72). These processes are driven by gravity, because melts and fluids are usually less dense than crystals, and by shear strain related to tectonics and larger scale movements of magma through the mush. Melt can migrate within a mush and may promote diffusive or convective exchanges between the melt in the mush and neighbouring magma chambers. Bubble growth within a mush (induced by crystallization and/or by decompression) can drive melt flow through the mush via conservation of mass (filter pressing). The physical role of volatiles may be underestimated. Even dry, mafic magma intrusions may have substantial CO₂ that is petrologically cryptic (e.g. has little effect on igneous phase relations), but physically important. It has been argued that many silicate melts can contain significantly higher dissolved CO₂ at high pressure than would be inferred from, for example, analysis of melt inclusions (73). Bubbles growing within mush can also produce channels by reactive flow (74, 75).

There is a profound transition in rheological properties between magma and mush over a narrow region of crystal content (45,47,66, 76). Mush rheology is largely controlled by deformation of the crystalline framework whereas in magma, rheology is largely controlled by deformation of the melt. The transition from mush to magma and vice versa can take place due to small changes in melt content or strain rate (at constant melt fraction) and typically involves several orders of magnitude change of viscosity (49,76). We term this the MMT (mush magma transition; Figure 3). The crystal content in the region of the MMT depends on crystal size, shape and shear rate and so can vary considerably, but is typically ~ 50 to 70%. From a fluid dynamic perspective the MMT can be regarded as analogous to a transition between solid (mush) and liquid (magma), although this is somewhat misleading since mush can deform as viscous material while the MMT temperature is typically much lower than the liquidus temperature.

Subsolidus rocks constitute the third domain of magmatic systems (Figure 3a). They contain no igneous melt by definition, but likely contain fluids, whose origin may be either magmatic (e.g. hydrothermal solutions or hypersaline brines) or external (e.g. meteoric water). Subsolidus rocks include both older country rocks into which the igneous system is emplaced and plutonic rocks formed from the igneous system itself. The latter may form when magma falls below the solidus temperature by either cooling or degassing or by highly efficient extraction of melt to form refractory rocks with very high solidus temperatures, as exemplified by adcumulate rocks. Much is known about the rheology of subsolidus rocks (54) in the context of crustal deformation due to tectonic processes. Rocks in the vicinity of or within active igneous systems will commonly be at high temperature. Important controls on the rheology of hot subsolidus rocks include mineralogy, presence or absence of small amounts of water (wet or dry), grain size and strain rate. Effective viscosities of many wet

crustal rocks just below the solidus fall in the range 10^{17} to 10^{19} Pa s; importantly wet conditions are likely prevailing in many igneous systems where solute-bearing H_2O is derived both from magmas and metamorphic reactions. The chemical reactivity of intergranular fluids may also play an important, but neglected, role in chemical transport and textural development in sub-solidus rocks. High-temperature, hydrothermal alteration of granites, for example by hydrolysis of feldspars at around $550^{\circ}C$, is one manifestation of such reactive flow.

Multi-component magmatic volatile phases (supercritical fluids or low-density vapours) can exist in all three magmatic system domains. They have very low viscosities (10^{-2} to 10^{-5} Pa s) and generate large buoyancies due to their much lower density than magmas and mushes. Volatiles may be multiphase in character; for example $NaCl-H_2O$ supercritical fluids can undergo phase separation upon decompression into relatively dense, high-salinity brines and low-density, low-salinity vapours (77). The low viscosity of fluids and vapours means that they can separate from magmas and mushes and invade sub-solidus host rocks.

Formation of magma chambers and reservoirs

Three main ideas have emerged on how magma chambers and magma reservoirs form; namely incremental intrusion, remobilisation and melt segregation from a mush. In reality these mechanisms can overlap and operate in harmony, but it is useful to distinguish them for clarity of discussion.

Incremental intrusion

It is now widely accepted that many intrusions are formed over a protracted length of time and show evidence of incremental growth often involving many repeated intrusive events (78-83). Formation of single event intrusions, such as sills, are of course recognised, but here our focus is on how to form magma reservoirs and chambers with significant magma volumes that can supply volcanoes and form plutonic rocks. In the crust the main issue is one of heat transfer and the basic condition for forming both a magma reservoir and a chamber is to advect more heat into a zone of magma intrusion than is lost by conduction, occasionally augmented by hydrothermal convection in the overlying country rocks. A reservoir is formed when persistent melt develops above the MCT, whereas a magma chamber forms when the melt fraction in a region exceeds the MMT.

The principles and findings of incremental growth models, as well as the geological and geochronological evidence for this mechanism, have been reviewed by Annen et al. (80). The main controlling factors are rate of input, ambient host rock temperature and duration. Key findings are summarised briefly here and typical results (82) shown in Figure 4. The basic model is that, if the increments of growth are quite small then early increments completely solidify. However, if the growth rate is sufficiently high then heat accumulates and temperatures increase with time within the intrusion zone. After an amount of time that depends on a number of factors, including the local geotherm, magma temperature and emplacement rate, intrusion depth, any leakage due to volcanism, intrusion geometry and thermal properties, the temperatures become high enough to form a mush, but melt fraction remains below the MMT temperature. The time required to develop a persistent melt is called the incubation time. As melt fraction increases further the MMT condition is exceeded (the activation time) and a magma chamber can grow. Emplacement rate provides

the strongest control on magma reservoir formation and a minimum rate is needed below which the incubation time is never reached (80, 81).

Some important more general concepts have emerged out of incremental growth models. In particular it is easier to form magma chambers deeper in the crust where host rocks are hotter (35,81-83). Evidence drawn from geology, geophysics, geochemistry and petrology has led to what is probably a consensus view that much magma processing and chemical differentiation occurs in the lower and middle crust, leading to the ideas of MASH processes and hot zones (35,84,85). Further, inferred average magma fluxes often seem too low to allow upper crustal chambers to form, leading to the inference that ascent of magma from deeper parts of the magmatic system to form shallow magma chambers must be episodic and associated with relatively short-lived episodes of high magma flux (86). We will discuss this matter later when attempting to integrate evidence and models.

One of the main issues for magma chamber formation by intrusion is the space problem. Single intrusions into sub-solidus rock in the elastic regime (sills and dykes) have very high aspect ratios (width to thickness), dictated by elastic laws of deformation (87). This mechanism is supported by geophysical as well as geological observations (88,89). However, many intrusions and inferred magma chambers have much smaller aspect ratios than estimated from elastic theory. Assimilation can only make a minor contribution to the space problem, whereas roof lifting is likely only significant in the very shallow crust (e.g. laccoliths). Stoping of roof rock blocks is similarly unlikely to provide a solution to the space problem (90). Thus a ductile mechanism is indicated in which host rocks are deformed to make space. Given that the crust within which magmatic systems develop will have strong temperature and rheological gradients, models in which the intrusion grows downwards into more easily deformable host rock are attractive (91-95). The requirement for ductile deformation to make the space to form magma chamber places additional constraints on the required growth rates. If growth rate is too fast then conditions can develop for brittle failure of the chamber wall, perhaps leading to eruption (95-97). Thus growth rates cannot be too low (insufficient heat advection) or too high (deformation too fast).

Tectonic controls are also important. In general tectonic stress fields influence the geometry of intrusions and magma reservoirs (e.g. 3, 87, 98, 99) that in turn feedback into the dynamics and overall evolution. There is a growing consensus from modelling, experiment and observation that extensional tectonics lead to frequent eruptions while compressive environments favour sill formation (e.g. 100, 101). Releasing bends on transcurrent faults may also facilitate shallow magma emplacement, by mitigating the space problem. In that case emplacement rates and tectonic deformation rates should be comparable. These are large topics in themselves and beyond the scope of this article to explore in detail.

Remobilisation

Here the essential concept, sometimes also termed 'defrosting' (102-104), is that the immobile domains of a magmatic system (mush, supersolidus rock and host rock) are converted into eruptible magma. Remobilisation models usually involve either heat transfer or fluid (water) fluxing, or both when new hydrous magma is emplaced below mush or host rock (e.g. 65,105) and may involve physical mixing of magma with mush (e.g. 105). When the new magma is hotter, the host rock or mush is heated well above the solidus, eventually

reaching the MMT and so transforming mush into magma (Figure 5). Crossing the MMT often leads to onset of convection and mixing. The detailed dynamics and character of the remobilised magma depend on the geometry of the new intrusions. For example heating from below by a sill-like body (105) generates a magma layer from remobilised host rock or mush, whereas progressive infiltration of host rocks by numerous dykes is conducive to hybridisation of the magmas (e.g. 107). Addition of heat and volatiles from new magma emplaced below the mush can lead to convection in the mush, facilitating its conversion into crystal-rich magma (105,108). Models of the remobilisation by hot basaltic recharge typically give rise to counterintuitive phenomena (105) such as melting of the magma layer margins and cooling and crystallization within the convecting interior of a growing magma layer (Figure 5)

Recent literature has focussed on mush remobilisation, especially in silicic systems (34, 65,105). Mushes with melt contents initially close to the MMT (typically 40%) require much less energy to remobilise by heating or by volatile fluxing than for sub-solidus host rocks or near-solidus mushes with low melt content (63, 109). Cold storage is a related but different concept, first articulated by Mahood (102). Here interpretation of geochronological data (principally on zircons) and trace element crystal zoning has led to the idea of protracted 'cold storage' of crystals at near-solidus or even sub-solidus temperatures followed by remobilisation of the mush by new, hot magma shortly before eruption (110,111). Evidence for cold storage has been presented as supporting thermal remobilisation (e.g. 111), but, as we discuss below, increasing the melt content across the MMT can occur by reactive melt infiltration accompanied by decompaction without either temperature increase or volatile fluxing.

There is clearly an overlap between remobilisation and models of incremental growth. If the latter involve episodic additions of new magma then each pulse may lead to limited remobilisation within the growing chamber until such point that a large volume of the reservoir lies above the MMT. One aspect of this model for rapid generation of large volumes of silicic magma is that commensurately high influx rates and large volumes of new magma are required to provide the heat and volatile source for remobilisation (86).

Magma chamber formation by reactive melt segregation in mush

Here the essential idea is to generate melt-rich layers within mush systems due to combined compaction and reactive melt flow (Figure 6). This is commonly depicted as synonymous with the process of extracting melt from mush to form melt-rich layers (e.g. 61, 112). However, these processes are not fully encapsulated in the notion of a one-way process, especially when reactions between melt and solid are introduced. In compacting systems some regions can start melt-poor and become more melt-rich by influx of melt (decompaction); under these conditions reactions between melt and crystals can generate melt-rich regions in which the MMT is exceeded. Reactive infiltration of melts enriched in low melting point components increases melt fraction by modifying the local bulk composition (113). Such processes enable the MMT to be crossed, for example, without any increase in temperature; for the case of reactive flow temperature may even decrease as the melt content increases due to latent heat effects (72,113).

In many situations buoyant melt moves upwards and the mush densifies as melt is extracted. By definition melt segregation processes can only take place above the solidus. Different physical mechanisms have been invoked to expel melt and concentrate crystals. In filter pressing (a variety of compaction) the solid grains (crystals) are rearranged but not necessarily deformed. Filter pressing can be enhanced in the presence of exsolved gas (114, 115) and can reduce porosity significantly (typically 1-20%), but not to negligible amounts, as evidently has happened in adcumulate rocks. The most widely invoked process for complete melt segregation is compaction via viscous deformation of the crystals. We re-evaluate compaction in the light of developments since McKenzie (61) and consider the challenge to the compaction hypothesis developed by Holness et al. (8).

McKenzie (61) outlined the principles of compaction of partially molten igneous systems to generate melt-rich regions. In a system of thickness h starting with a uniform distribution of melt at a fixed porosity the compaction length, δ_c , can be identified as follows:

$$\delta_c = [\mu^{-1}(\xi + 4/3\eta) k_\phi]^{1/2} \quad (1)$$

where μ is the melt viscosity, ξ is the bulk viscosity, η is the shear viscosity and k_ϕ is the permeability. A time scale, τ_c , can be defined as the time for the base of the compacting layer to reduce its porosity by a factor of $1/e$. For a layer with the thickness of the compaction length scale this time scale is given by:

$$\tau_c = \{\mu \Phi \delta_c\} / \{k_\phi (1 - \Phi)^2 \Delta \rho g\} \quad (2)$$

where Φ is the porosity (melt fraction), g is gravitational acceleration and $\Delta \rho$ is the density contrast between the melt and crystalline matrix. A minimum time scale occurs when $h = \delta_c$; for $h < \delta_c$ the time scale increases in proportion to layer thickness while for $h > \delta_c$ the system evolves into a series of porosity waves with wavelengths of approximately δ_c .

Bulk viscosity is a critical parameter in compaction problems and was held constant at a value of 10^{18} Pa s in the original calculations presented by McKenzie (61). However, empirical and theoretical studies indicate that the bulk viscosity varies greatly with melt fraction and that at high melt fractions it is much lower. At very low melt fractions the bulk viscosity is about three orders of magnitude larger than the shear viscosity; they approach comparable values at high melt fractions according to theoretical models (62). Bulk viscosity values of 10^{15} Pa s or even lower are indicated by studies of porosity reduction in cumulates from mafic layered intrusions (116, 117). For a fixed melt viscosity the variation of bulk viscosity with melt fraction indicates decreasing length scales and time scales for compaction with higher initial melt fractions. We have re-calculated length scales, melt velocities and time scales for melt viscosities of 10^5 Pa s (wet rhyolite), 3×10^2 Pa s (andesite) and 1 Pa s (hot basalt) taking account of the variation of bulk viscosity. Representative parameters are listed in Table 1 and the results are displayed in Table 2 using equations 1 and 2 above and equation 2 from McKenzie (61). Our re-assessment gives generally shorter time scales than McKenzie's. A surprising result is that the length scale and time scales are not strongly correlated with porosity. McKenzie (61), using a constant bulk viscosity, found strong positive correlations of compaction (length and time scale) with increasing porosity, whereas we derive weak negative correlations with increasing porosity.

Since these ideas were developed there has been much modelling research on compaction-driven segregation, largely in relation to partial melting and extraction of melts from the mantle (60, 118) and behaviour of metamorphic fluids in the deep crust (119,120). Research increasingly focuses on crustal, rather than mantle, igneous systems (72, 85, 113, 121, 122). Discoveries include: the instability of melt waves in 2 dimensional models with formation of vertical flows of melt along high permeability pathways; reactions between melts and crystalline matrix (reactive flow) including latent heat effects as matrix is either dissolved or precipitated due to reactions; formation of melt-rich regions which have compositions of small degree melts; that the role of an exsolved volatile phase in some circumstances inhibiting melt crystal segregation; and the important role of the solidus isotherm in limiting the upward movement of melts.

The last effect is particularly important for understanding formation of melt-rich bodies at the top of a mush system beneath a solid rock and is captured in the models presented by Jackson et al. (113, 121). In 1-D models melt moves through mush forming melt-rich waves and residual melt accumulates under rock at the top of the mush. This melt layer will grow with time unless it is erupted or transported to a higher level in the crust. The melts are typically highly differentiated and low temperature. The process explains well the development of a zone of crystal-poor magma overlying crystal-rich magma, mush or cumulates with low melt fractions in both basaltic systems (e.g. 26) and silicic systems (33). Volatiles play an important role in such processes because of the effect of dissolved H₂O on magma solidus temperature. Ascent of water-saturated magma below a certain threshold pressure where the solidus temperature increases sharply leads to crystallisation (123). This problem is overcome if melts are water-undersaturated, for example by the presence of dissolved CO₂. In these cases the effect of volatile-exsolution serves to increase the activity of H₂O in the melt, reducing the solidus temperature and preventing freezing (73).

The compaction hypothesis has been challenged principally on the basis of textural observations of cumulate rocks (8) and granites (124). Holness et al. (8) point out that there is no evidence for internal deformation of primocrysts in adcumulate rocks from layered intrusions, notably the Skaergaard, Greenland. They find evidence for crystal deformation in some very large layered intrusions, like the Bushveld, South Africa, but argue that the observed amount of deformation cannot have reduced significantly the porosity. We agree with their interpretations and reasoning, but we are less convinced in the arguments against compaction by a dissolution-precipitation mechanism. Conceptual models for textural development due to dissolution-precipitation have been developed for tectonic deformation of sub-solidus crustal rocks in the presence of small amounts of fluid (8). Here dissolution occurs along grain boundaries normal to σ_1 while extension in the σ_3 direction creates spaces in which crystals can grow. However, there are some significant differences in a mush, especially at higher porosities, that suggest the behaviour will be different. First, stress distributions in the vertical (loading direction) will be much more heterogeneous in magnitude and direction, as they depend on local geometry of crystal contacts. Second, compaction is unidirectional and we infer precipitation will take place within pores, likely along protruding free crystal faces. Third, melt is reactive with solids and there will be continuous local changes of phase proportions and bulk compositions as melt migrates during compaction (e.g. 72, 113) redistributing mass locally and enabling deformation. We

acknowledge that there are challenges in distinguishing crystal growth via this mechanism from that due to other mechanisms such as compositional convective or diffusive exchanges between the mush and an overlying magma chamber. However, textural and chemical evidence for dissolution-reprecipitation processes in mafic rocks from the lower oceanic crust are compelling (71, 125,126). Moreover, reactive flow can create high melt fraction layers with relatively little compaction of the underlying mush (of order 10% reduction in thickness) that may be difficult to detect in preserved textures (113).

Finally we note that reactive flow in mushes offers an alternative explanation of cold storage to remobilisation by heating or gas fluxing. Porosity waves are an important feature of compacting systems and melt fraction can increase as well as decrease. Decompaction and reactive flow involves flooding of melt into low melt fraction regions and is capable of disaggregating mush and crossing the MMT to form magma and thus disperse old crystals. Jackson et al. (113) show that magma chamber formation by this mechanism is an inevitable feature of hot zone systems formed by incremental intrusion with reactive flow. Their models do not, however, consider the effects of dissolved volatiles on melt phase relations and solidus temperatures, as discussed above.

Dynamic magma transport through magma reservoirs

Magmatic transport through the crust is a major aspect of magmatic systems and the subject of a vast literature. Indeed formation of magma chambers and reservoirs by the intrusive mechanisms discussed above is one part of the magma transport story. Transport through subsolidus host rocks occurs either by brittle mechanisms (e.g. sills, dykes and laccoliths) or by ductile mechanisms (e.g. diapirism and down-sagging). Much of the understanding of these processes relates to intrusions into cold rigid rocks in upper crustal regimes that are well below the solidus. However, magma chambers are commonly quite shallow relative to crustal thickness and magma input from deeper levels is implicit in intrusive and defrosting models of magma chamber formation. Models of magma transport through dykes within deeper and hotter more ductile parts of the crust are either implicit or explicit (e.g.127). Dyke transport through the entire crust is supported by geophysical (e.g. 128, 129), geological (130), and petrological (131) evidence in some volcanic eruptions. Much less attention has been paid to magma transport within mushes. The much lower effective viscosity and mechanical weakness of mush compared to sub-solidus rocks means that transport in mushes could be fast and might enable large magma chambers to form rapidly (e.g.36,38). This topic is the focus of this section. There are several different circumstances in igneous systems where a buoyant layer can form within a mush system. Two main situations occur to us: emplacement of a new layer of magma at the base of or within mush (replenishment), and formation of melt-rich layers within a mush due to compaction and melt flow. We consider each in turn.

Mush reservoir replenishment

This situation develops in multistage magmatic systems where magma generated at one level is transported to a higher level and encounters a mush zone. The new magma may be emplaced within or at the base of a mush system at a level of neutral buoyancy or where there is a significant change in mush rheology associated with the transition across the MCT. We have already discussed the case where heat and volatiles from the new magma transform the overlying mush into magma (thus forming a magma chamber in situ).

However, it seems likely that there will be many cases where the mush is either not remobilised or the remobilised zone is only a small fraction of the total mush thickness. The variety of cases (depicted in Figure 8) envisaged is similar to those related to the replenishment of magma chambers, extensively investigated in the 1980's (e.g. 18, 21, 132). The difference is that in magma chamber replenishment models the viscosity contrast between the layers is relatively small and under buoyancy-driven, unstable conditions the magmas can mix readily together. In the case of magma emplaced into a mush the viscosity contrast between magma and mush will typically be larger by many orders of magnitude.

Figure 8 shows some likely common scenarios for a layer of denser and hotter magma emplaced within or at the base of a mush. The magma layer cools by convection or conduction and transfers heat (\pm volatiles) and melt to the overlying mush. Some end member cases can be envisaged, their realisation depending on many different factors. In one situation (1 and 2 in Figure 7), cooling of the emplaced layer results in formation of crystals which settle out (or accrete to the magma layer floor). The density of the remaining magma decreases and eventually becomes buoyant resulting in ascent as a dyke (case 1) or a plume (case 2). In case 3 crystals remain in suspension but the magma layer develops buoyancy due to some volatile exsolution, becomes unstable and forms a plume. A variant of these scenarios is when the melt fraction in the mush above the new magma layer increases sufficiently to cross the MMT and so is transformed into crystal-rich magma. Such an overlying remobilised mush layer might either go unstable leaving the replenished magma behind (case 4) or could mix with the underlying replenished magma to form a stable layer (case 5) or become unstable (case 6). If magma is emplaced within the mush at a level of neutral buoyancy then the same scenarios can be anticipated, but an additional factor might be development of negative buoyancy in which the emplaced layer becomes denser with time or forms dense cumulates at its base, and so can form plumes that sink into underlying mush (e.g. 94).

Compaction generated layers

Models of compaction of partially molten systems show development of instabilities in the form of waves of high porosity (10, 113), which are inherently unstable. These porosity waves may lead to accumulation of significant melt layers beneath zones of low or no permeability. Such accumulations can occur at the top a magma reservoir where the solidus defines zero melt permeability, but can also occur within compacting regions due to reactive flow and geological heterogeneity (72). In two dimensional models of compaction the developing high porosity waves may go unstable (due to buoyancy and reactive flow) and form vertical melt-rich fingers rising through the system (60, 119, 133).

There are also mechanisms to create multiple melt-rich regions within a compacting mush system. Compaction models typically assume rather homogenous properties of the compacting system but real geological systems are likely quite heterogeneous with vertical and lateral variations in compositions; layered intrusions are an obvious example of this heterogeneity. Thus there can be more refractory zones beneath which melt can accumulate. In addition, for the case of reactive flow Solano et al. (72) have shown that melt-rich layers can form by melt moving from one composition rock to another and then reacting with the new rock layer to generate melt. In this case melting is achieved not by increasing temperature at fixed composition, but by changing composition at fixed

temperature. Such a situation is likely to be diagnostic of percolative reactive flow in both simple (72) and natural systems where melt fractions and melting temperatures are very sensitive to composition.

Instability of buoyant layers within mushes

Although extremely viscous, the deformability of mush means that a layer of buoyant magma within a mush is inherently unstable. Such instability is consistent with geochronological and petrological evidence for the episodic nature of volcanism and magma intrusion (80, 133). With respect to the formation of magma chambers and lenses there must be conditions for which magma can accumulate in a layer faster than gravitational instabilities remove it by buoyant ascent of magma blobs through the mush. Numerical investigations by Keller et al. (133) of a buoyant source emplaced at the base of a high viscosity domain indicate that diapirism will be a dominant instability and transport mechanism within mushes.

An investigation of this situation has been presented by Seropian et al. (135) based on Rayleigh-Taylor instability theory developed by Whitehead and Luther (136) and building on experiments and further theory development by Bremond d'Ars et al. (137). Seropian et al. (135) draw attention to the issue that, in many magma systems, the width-of a buoyant magma layer will be much less than the theoretical wavelength of the fastest growing instability based on theory for laterally-infinite layers; they also show that the volume of magma that can accumulate is very sensitive to the mush rheology.

A useful starting point is the case of a buoyant layer that is instantly emplaced within an igneous mush system. This case is valid provided the time scale of layer formation (e.g. by mush compaction, gas-filter pressing or magma replenishment) is much shorter than the time scale of the instability, which can be queried *a posteriori*. In the case of a magma intrusion that is emplaced at its level of neutral buoyancy in a mush, post-emplacement processes, such as vesiculation or crystal settling with or without accompanying magma convection, can generate a buoyant layer. In the latter case the relevant time scale for layer formation is related to cooling, crystallization and segregation of melt from crystals, and reduction in layer density due to volatile exsolution.

According to Rayleigh-Taylor instability theory, perturbations in the thickness of the buoyant layer of all wavelengths will grow but at different rates, eventually generating rising blobs of the buoyant magma with a spacing corresponding to the fastest-growing wavelength, λ_c . For a magma layer of thickness h , and shear viscosity μ_1 , below a denser and much thicker layer of mush with shear viscosity μ_2 ($\mu_2 > \mu_1$), the fastest-growing wavelength is:

$$\lambda_{RT} = 4.36h(\mu_2/\mu_1)^{1/3} , \quad (3)$$

that grows at rate:

$$n_{RT} = 0.232\Delta\rho gh \mu_2^{-1}(\mu_2/\mu_1)^{1/3} , \quad (4)$$

where g is gravity and $\Delta\rho$ is the density difference between the layer and the overlying mush (136). The characteristic timescale for the growth of the instability (i.e. the time for the amplitude to increase by a factor of e) is $\tau_{RT} = 1/n_{RT}$. Notably, this theory is for layers that are laterally infinite. However, due to very large viscosity contrasts between low-crystallinity magma layers and mushes, the predicted length scales for the fastest growing instability typically cannot be attained because they greatly exceed the likely horizontal dimensions of magmatic systems (135). For example using a representative value of $h = 10$ m, $\Delta\rho = 300$ kg/m³ and $\mu_2 = 10^{14}$ Pa s results in λ_{RT} values of ~ 2000 km, 300 km and 40 km for μ_1 of 1, 300 and 10^5 Pa s, respectively. Thus it is more appropriate to investigate confined instabilities where the largest possible wavelength is the lateral extent of the magma layer, L .

When the system is confined ($L < \lambda_{RT}$) the instability develops as a single broad protrusion that grows into a spheroidal pocket of magma that rises through the mush (135). The wavelength of the instability is L (the maximum length scale available) and the growth rate is reduced by a factor of L/λ_{RT} compared to the unconfined case (Eq. 4), and can be expressed as:

$$n_{\text{confined}} = \Delta\rho g L / (6\pi\mu_2). \quad (5)$$

Notably, the growth rate for the confined case does not depend on the thickness or viscosity of the magma layer, as it does for the unconfined case (Eq. 3), but does depend on the viscosity of the mush, μ_2 . Therefore the characteristic timescale for development of the instability ($1/n_{\text{confined}}$) depends largely on the viscosity of the mush, which can vary by several orders of magnitude. A consequence is that the mush rheology also strongly affects the maximum thickness (and volume) of magma that can accumulate prior to the onset of instability for a given rate of thickening of the magma layer (dh/dt). The magma layer can only thicken if magma is added (by intrusion or by mush compaction etc.) faster than it is removed by the Rayleigh-Taylor instability. Modifying the criteria of de Bremond d'Ars et al (137) to take into account laterally-confined instability, the maximum possible thickness of the magma layer is:

$$t_{\text{max}} = n_{\text{confined}} dh/dt. \quad (6).$$

Considering mush viscosities of 10^{13} - 10^{17} Pa s, Seropian et al. (135) calculate (Figure 8) a large range of timescales (10^{-1} - 10^4 years) and magma volumes (10^5 - 10^{12} m³). These time scales and magma volumes bracket most volcanic activity. Thus the instability of magma layers within mush systems can account for episodic magma transport through mush domains of magma reservoirs. In turn these instabilities can help explain the episodic accumulation of magma at high levels in magmatic systems to form shallow magma chambers and intrusions, and which eventually feed volcanic eruptions (138).

Finally we note that during diapir ascent further melt-crystal-fluid segregation can take place within the diapir by crystal settling or bubble rise and, for crystal-rich systems, by compaction (133).

Fast transport through mushes

Diapirism is not the only transport mechanism for melt and fluids through mush. The numerical simulations of Keller et al (133), although using parameters not directly applicable to mush systems, qualitatively show three forms of transport: diapirism, development of vertical melt-rich channels (or chimneys) and development of tensile fractures. Under fast rates of deformation a mush may fracture and allow very fast rates of magma ascent. The characteristic features of mafic inclusions (enclaves) in both plutonic and volcanic rocks can be explained by propagation of mafic magma dykes through mush followed by disaggregation (e.g. 140-142). In the case of volatiles Christopher et al. (143) interpret geophysical and geochemical data on bursts of gas flux at the Soufriere Hills Volcano, Montserrat as evidence for fluid-filled fractures propagating through at least 5 km thickness of mush.

Mush re-organisation

An emerging idea, building on concepts discussed in the preceding sections, is that magma chambers can form rapidly within mush systems by instability of multiple magma layers which merge rapidly together at the top of the system (36, 38). These ideas have developed to explain the evidence that very large silicic magma chambers that are associated with caldera-forming eruptions can be formed in only a few decades or even less (30,39, 144). A similar concept has been applied to explain very high volcanic gas emissions during and following some major eruptions where fluids trapped within a mush system are rapidly liberated (143).

Thus we can identify another, highly dynamic way of generating a large magma chamber within a magmatic system by rapid amalgamation of multiple magma and fluid layers within the mush domain. In this case the effects of decompression may be particularly significant with upward transport of magma (fluids). Ascent can result in crystallization if the magma is water-saturated (146). Large overpressures (pressures in excess of lithostatic) can develop as decompression-induced volatile exsolution occurs, with or without attendant crystallisation in the case of water-saturated melts (147). A cascading process can be envisaged where, as unstable magma layers merge, their volume increases and buoyant forces increase so that transport through the mush accelerates. This concept also becomes an attractive idea for the triggering of volcanic eruptions and explaining the highly episodic character of volcanism. It also explains the petrological evidence from large volume eruptions for amalgamation of disparate, but cogenetic, pockets of melt prior to or even during eruption (e.g. 38,39,148,149). Mush development and multiple magma layer formation is a very slow process that takes place during periods of volcano dormancy, while instability and magma transport within a mush is a very rapid process leading to a period of volcanic unrest and sometimes eruption.

A major difference between this idea and the well-established idea of magma chamber recharge to accumulate and pressurise magma in the upper crust and trigger volcanic eruptions is that pressure increases are not caused by the transport of the incompressible components of magma (melt and suspended crystals) because upward magma movement is compensated by mush moving downwards. In a mush re-organisation scenario, pressure and volume changes in the developing shallow magma chamber can only be caused by decompression and exsolution of the compressible volatiles. The implications for understanding volcanism have been explored in Sparks and Cashman (149). Of particular

note volume changes that result in surface deformation need not be directly related to magma recharge volumes. Moreover, magma accumulation does not require significant, long-term mush heating in order to occur, consistent with the concept of cold storage, whereby volcanic crystals display limited evidence of pre-eruptive heating (e.g. 150,151).

The emerging paradigm

This paper has largely been concerned with understanding how the physical properties of magmatic systems lead to a very wide range of dynamics. We have defined rheologically different domains that depend on the presence and fraction of melt (and fluids). We have focussed on different mechanisms by which melt-rich bodies (magma chambers) can form and on the dynamics of magma transport within magma reservoirs. Here we briefly set these topics in a wider context of igneous processes, including differentiation, magma transport, magma system evolution and magma system structure.

Crustal magmatic systems develop in different tectonic settings by the fluxing of magma into and through the crust, leading to formation of intrusive rocks and volcanism. We can envisage different stages of thermal maturity in which early magma fluxes warm the crust but freeze, whereas later fluxes enable substantial magma reservoirs to form (Figure 9). These magma reservoirs (or hot zones) may form as disconnected bodies but, as more and more magma is added, may become increasingly interconnected. Cooling and crystallization of the magmas leads to separation of melts from crystals by diverse mechanisms and differentiation, while buoyancy leads to instability and transfer to higher levels of the developing system, leaving behind refractory crystal residue. Remelting of older rocks and intrusions is inevitable, except where country rocks are unusually refractory, and can contribute additional magma. The chemical character of crustal melt will differ from that of the input melt, leading to production of magmas with mixed mantle and crustal sources, a feature widespread in continental magmatism (35).

Based on understanding of the processes some generalisations about crustal magmatic system structure and evolution can be made. In any kind of crust a density, thermal and mechanical structure exists prior to any magma injection (Figure 9). With depth the crust gets denser and hotter, while first becoming stronger with depth in the brittle upper crust, but then switching to becoming weaker with depth in the ductile middle and lower crust. These pre-existing gradients become modified by magma fluxing but the fundamental structure remains and provides a first-order control on how an igneous system evolves. Of course the modifications to these gradients by magma fluxing, intrusion and tectonic regime can be marked with increasing temperatures shifting the brittle to ductile transition to shallower depths. Also the development of domains containing melt in the crustal column greatly reduces viscosity, while igneous differentiation processes change the density gradients.

Crustal structure and property gradients play a key role in where magmas tend to stall. Density gradients lead to magmas of more evolved composition tending to stall at higher levels in the crust. Marked density changes, such as the mid-crustal Conrad discontinuity or the geophysical Moho, can provide preferential sites for magma intrusion. Crustal strength reaches a maximum in the middle crust and crustal viscosity also decreases with depth. Thus

the middle crust is a region where conditions for magma stalling are particularly auspicious due to marked increase in rigidity (152) as well as density discontinuities. At the mature stage of a system (Figure 9) another factor favouring middle crustal melt accumulation is the solidus. Melt fluxing through regions of mush accumulate below the depth of the solidus temperature (113). The vagaries of geology together with tectonic stresses will also provide places for preferential stalling. It is when magma stalls that the processes we have discussed can begin.

Thermal controls are of great significance and here, following the hot zone concept (35), conditions favourable to developing a substantial magma reservoir increase with depth. Thus most chemical differentiation takes place in the lower and middle crust while the upper crust, being initially cold only forms substantial magma reservoirs because of transient high magma fluxes or prolonged periods of deeper magma intrusion and fluxing. Thermal structure is important for another reason. In the hot lower crust basaltic reservoirs that form have rapid segregation of melt and crystals due to the relatively low melt viscosity leading to refractory ultramafic and mafic cumulates; the residual more evolved melts are buoyant and ascend to shallower depths. For compaction-driven segregation the time scales for segregation and escape are geologically rather short (Table 2) so we do not expect much melt to accumulate in the lower crust long term. This expectation is consistent with geophysical imaging which infers rather little melt in the lower crustal parts of large magmatic systems (e.g. 153-156). More evolved melts can rise to higher levels in the system and stall, where the lower temperatures lead to further differentiation. Indeed the geotherm is likely to determine the amount of differentiation. Thus shallower magma reservoirs can form, from which more evolved and buoyant melts can be extracted. However, the higher melt viscosities mean that the processes of differentiation and melt segregation (Table 2) are much slower so distributed melt can be present for much longer periods of time, which might explain why interpretations of geophysical images suggest higher melt fractions in the middle regions of the crust (153-156).

An interesting corollary of the above scenarios is that melts that emerge from the tops of mid-crustal reservoirs will tend to have compositions controlled (or “buffered”) by the low-variance mineralogy of the surrounding, volumetrically dominant, crystal mush. Once these melts ascend to shallower levels and undergo decompression-driven crystallisation, or cooling-driven crystallisation in shallow plutons, their mineral assemblages may differ from those of the mid-crustal mushes from which they emerge. This allows for ‘cryptic fractionation’ of minerals that are not present in erupted magmas, such as amphibole in many intermediate arc magmas (157) or clinopyroxene in the case of mid-ocean ridge basalts (158). The recent proposal (159), based on experimental petrology, that arc dacites from Mount St. Helens volcano represent the products of peritectic crystallisation of melts in the low-variance assemblage amphibole + clinopyroxene + orthopyroxene + plagioclase + oxides in the mid to deep Cascades crust (700-900 MPa) lends further support to these ideas. Evidently mid-crustal mushy systems may lead to quite different patterns of igneous differentiation than those predicated on simple liquid lines of descent of mantle-derived parental magmas.

The upper crust seems to be predominantly a region of magma accumulation rather than differentiation. Magmas stall in the upper crust and require high flux rates to maintain a

magma chamber. Degassing, as well as cooling induce solidification, but with limited potential for the crystal-melt segregation process that drive differentiation. Annen et al. (35), for example, suggest that the chemical identity of a magma is defined by lower and mid-crustal processes, whereas the textural and mineralogical character of a magma are defined by the ascent and crystallisation path it takes through the shallow crust. For systems generating large, upper crustal silicic magma bodies igneous differentiation occurs over long time periods in the lower and middle crust hot zones; shallow chambers form rapidly as melt leaves these hot zones, acquiring great textural diversity, but little further chemical variety.

Final remarks

Understanding the formation and dynamics of magma reservoirs has developed over more than a hundred years with the first recognisably modern concepts emerging at the beginning of the 20th century. Magma chambers in particular have shaped thinking on many fundamental issues of crust formation, chemical differentiation, plutonism and volcanism. There have been many twists and turns and the focus of interest has shifted in different decades. As is usual in science new observations (notably from geophysics and geochronology) and new technical developments (notably analytical techniques with ever greater spatial resolution) have required new thinking. In the early 21st century we now are developing an understanding that provides considerable explanatory power by integration of many diverse observations, experiments and modelling. The emerging new paradigm is of course raising new research questions.

Magmatic systems are fundamentally divided into three principal domains: magma, mush and surrounding host rocks. Magma reservoirs are composed of the magma and mush domains and have three constituent phases with very different physical properties: melt, fluids and crystals. There can be transitions between the domains as heat and mass are transferred within the system leading to change in state. Rheology is a critical property and the fact that the viscosity of the domains and components magmatic systems varies by 25 orders of magnitude, from the very low viscosity of gas to the very high viscosity of the hot rocks that surround magma reservoirs, plays a fundamental role in shaping the dynamics of magmatic systems. This viscosity range helps explain the huge diversity in igneous phenomena and in the rates of igneous processes. It represents a substantial challenge to any modelling efforts.

Our review of the emerging ideas and themes on magma reservoirs enables a more nuanced view of magma chambers and how they form. It now seems clear that magmatic systems are predominantly mush, but magma chambers can form by a variety of mechanisms. Four main mechanisms can be recognised, although they are not mutually exclusive. In one mechanism magma is intruded into host rock at a rate that is sufficient to avoid cooling and solidification (80). In a second mechanism, sometimes known as defrosting, host rocks or mush are converted into a magma by heat or volatile transfer with magma recharge often invoked as a cause (65). In a third mechanism, a magma chamber forms by accumulation of melt (\pm entrained crystals and volatiles) by reactive flow through the mush (113). Here compaction is the most commonly invoked process, but there are others, such as volatile fluxing, compositional convection and mush shearing, that can lead to melt (or magmatic fluid) extraction. Moreover, changes in bulk composition in response to reactive flow may

play a more important role in controlling melt fraction than compaction. Broadly, time scales of extraction increase as melt viscosity increases. The fourth mechanism requires the first three mechanisms to form smaller magma layers due to recharge of new magma into the mush or formation of magma layers by extraction. These multiple layers become unstable and merge to form a larger magma body (38, 144). Whereas the first and third mechanisms suggest very slow processes, the second and fourth mechanism may be much faster. Understanding mechanisms by which disparate, cogenetic magma layers from within a single magmatic system can coalesce rapidly prior to or during eruptions represents a major avenue for future research.

On the larger scale, magmatic systems develop within and ultimately impose physical property gradients of density, temperature, strength and viscosity in the crust (Figure 9). These gradients, together with tectonic conditions, influence how magmas ascend, where magmas stall and how magma reservoirs develop to enable the mush and magma chamber processes discussed in this paper to take place.

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Figure Captions

Figure 1. Depiction of a classic magma chamber beneath a volcano in which the chamber is envisaged as a melt-dominated body recharged from below by new magma from depth and supplying volcanic eruptions.

Figure 2. Depiction of a thermally mature transcrustal magmatic system, where melt processing in a volumetrically dominant mush domain leads to formation of multiple magma chambers in the middle and upper crust. Diagram modified from Cashman et al. (36).

Figure 3. Diagram depicting physical and rheological properties of major domains within magmatic systems.

Figure 4. Examples of numerical models of silicic magma chamber formation by incremental growth (80). In this calculation, dacite magma is emplaced at a rate of 2.5 cm/year (a) 5 cm/yr (b), and 10 cm/year (c) into a crust with a geothermal gradient of 25°C/km. The models assume that each increment is emplaced below the last (overaccretion). The accumulated intrusions, shown by the dashed green contour, is 1 km on the right and 5 km on the left. Contours of melt content are shown. At low emplacement rate a small mush zone forms at 200 ky while at the high rate a large magma chamber has formed at 50 ky.

Figure 5. Example of a remobilisation model in which granitic rocks are heated, melted and converted into convecting magma by heating from below due to recharge of basaltic magma (after 105). Note the counterintuitive result that melting occurs at the boundary (roof) of the system while cooling and crystallization occurs in the convecting interior as the magma layer grows.

Figure 6. Numerical models to illustrate the development of a magma chamber by a combination of compaction and reactive flow. Results are shown after 1.51 My of steady incremental growth of a 4 km thick basalt intrusion. Each increment of growth is emplaced near the top of the intrusion (overaccretion). Evolved melt percolates through the evolving mush and accumulates at the top to form a magma chamber. Modified from Jackson et al. (113).

Figure 7. Cartoons of mush replenishment showing some variants of possible scenarios (see text and labels).

Figure 8. Time before the onset of instability and accumulated volume in the case of a linearly growing buoyant layer of silicic melt under (a) a melt-rich mush and (b) a near-solidus mush. The kinks in the lines in (a) for $dh/dt = 0; 1; \text{ and } 10 \text{ m/yr}$ correspond to a change from confined to unconfined instability regime with increasing melt layer width. There are no kinks in (b) because for the higher mush viscosity, Rayleigh-Taylor instabilities are confined for the full range of scenarios plotted (after 135).

Figure 9. Schematic profiles of density, strength, temperature and viscosity through the crust. A depth scale is not shown as the thickness of the crust varies greatly between

different tectonic settings (from 5 km at ridges to 70 km in continental arcs). Parameter values are indicative. Two contrasting end members are shown. In the upper panel crust is shown in its unperturbed state. The igneous system on the right is an immature system which might characterize either the early stages of magmatism or a low magma flux setting. The crustal properties are not much changed by small additions of igneous rock. The lower panel shows a mature system where the igneous activity has strongly perturbed the physical property structure of the crust due to advection of heat and mass to create a transcrustal magmatic system. The unperturbed property profiles are shown as dashed lines. The colour scheme indicates the transition from cold, strong, elastic brittle upper crust (blue) to hot, weak, ductile lower crust (orange to red as temperature increases).

Table 1. Parameters selected for compaction calculations shown in Table 2.

Porosity	0.01	0.03	0.1	0.3
Bulk viscosity Pa s	10^{18}	10^{17}	10^{16}	10^{15}
Permeability m^{-2}	3×10^{-14}	10^{-13}	3×10^{-13}	10^{-12}
Shear viscosity Pa s	10^{15}			
Density contrast kg/m^3	300			

Table 2. Calculations of compaction length, compaction time and melt viscosity for three different viscosities corresponding to basalt (1 Pa s), andesite (300 Pa s) and rhyolite (10^5 Pa s). The columns are for a range of initial porosities from 0.01 to 0.3. Parameters are listed in Table 1.

	Melt Viscosity Pa s	0.01	0.03	0.1	0.3
Length (metres)	1	173	100	58	48
	3×10^2	10	5.8	3.3	2.7
	10^5	0.54	0.31	0.18	0.15
Melt velocity (m yr^{-1})	1	0.27	0.29	0.25	0.21
	3×10^2	9.0×10^{-4}	9.8×10^{-4}	8.2×10^{-4}	7.1×10^{-4}
	10^5	2.7×10^{-6}	2.9×10^{-6}	2.5×10^{-6}	2.1×10^{-6}
Time (years)	1	660	349	267	98
	3×10^2	1.1×10^4	6115	4468	5474
	10^5	*	*	8.1×10^4	1.4×10^5

* below 7% melt

Figure 1.

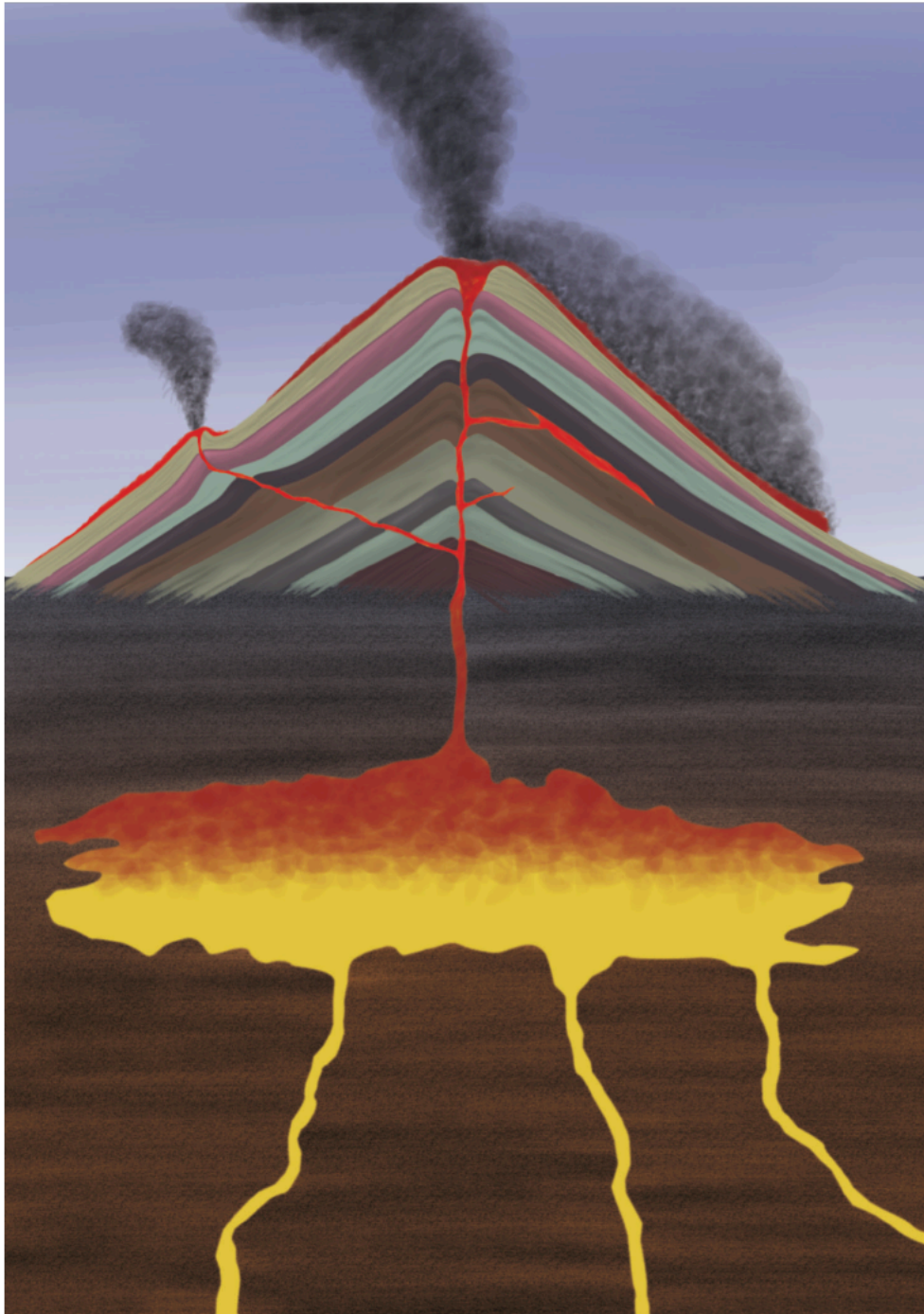


Figure 2.

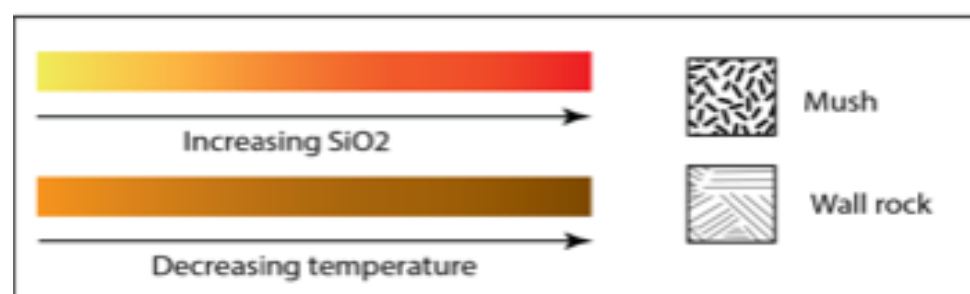
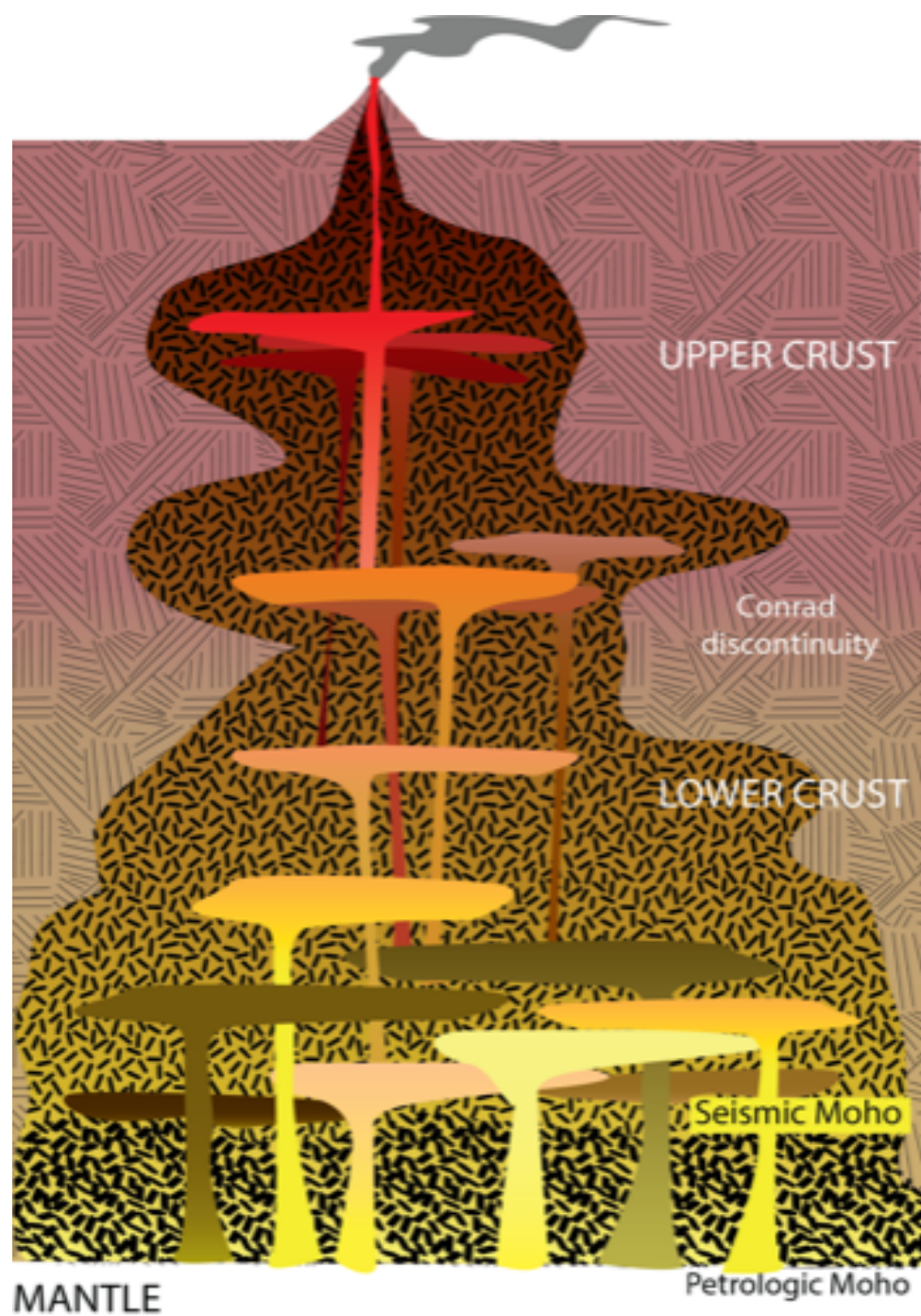
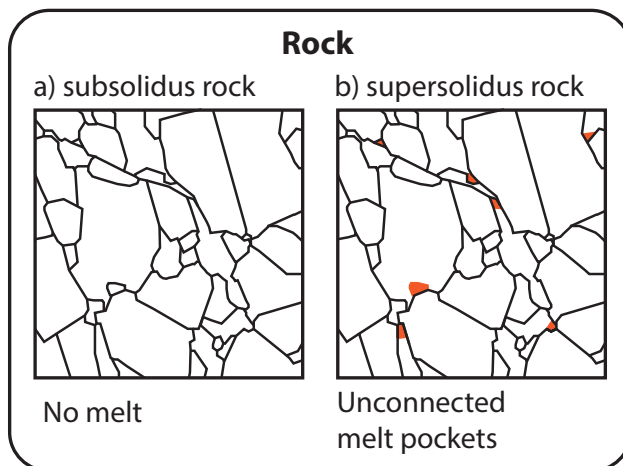
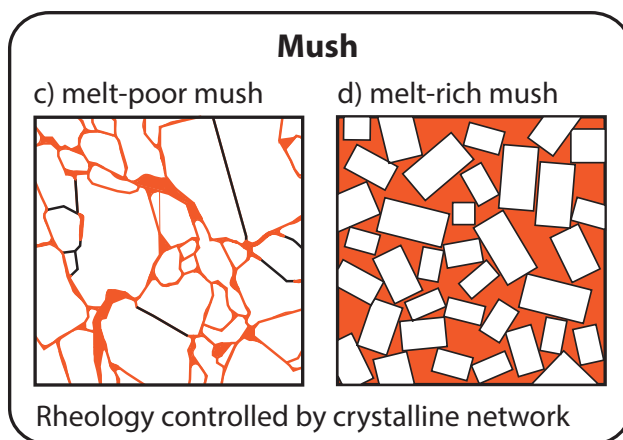


Figure 3.



Melt Connectivity Transition
gabbro: ~ 1% melt; granite: ~7% melt
Large reduction in effective viscosity and strength



Mush-Magma Transition
Typically 30-50% melt, depends on crystal size distribution and shape, and shear rate

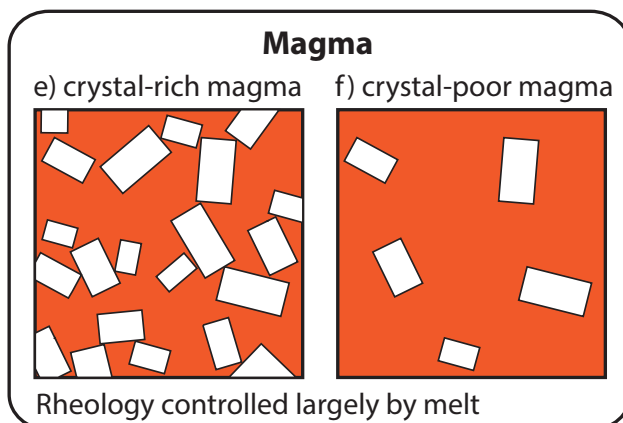


Figure 4.

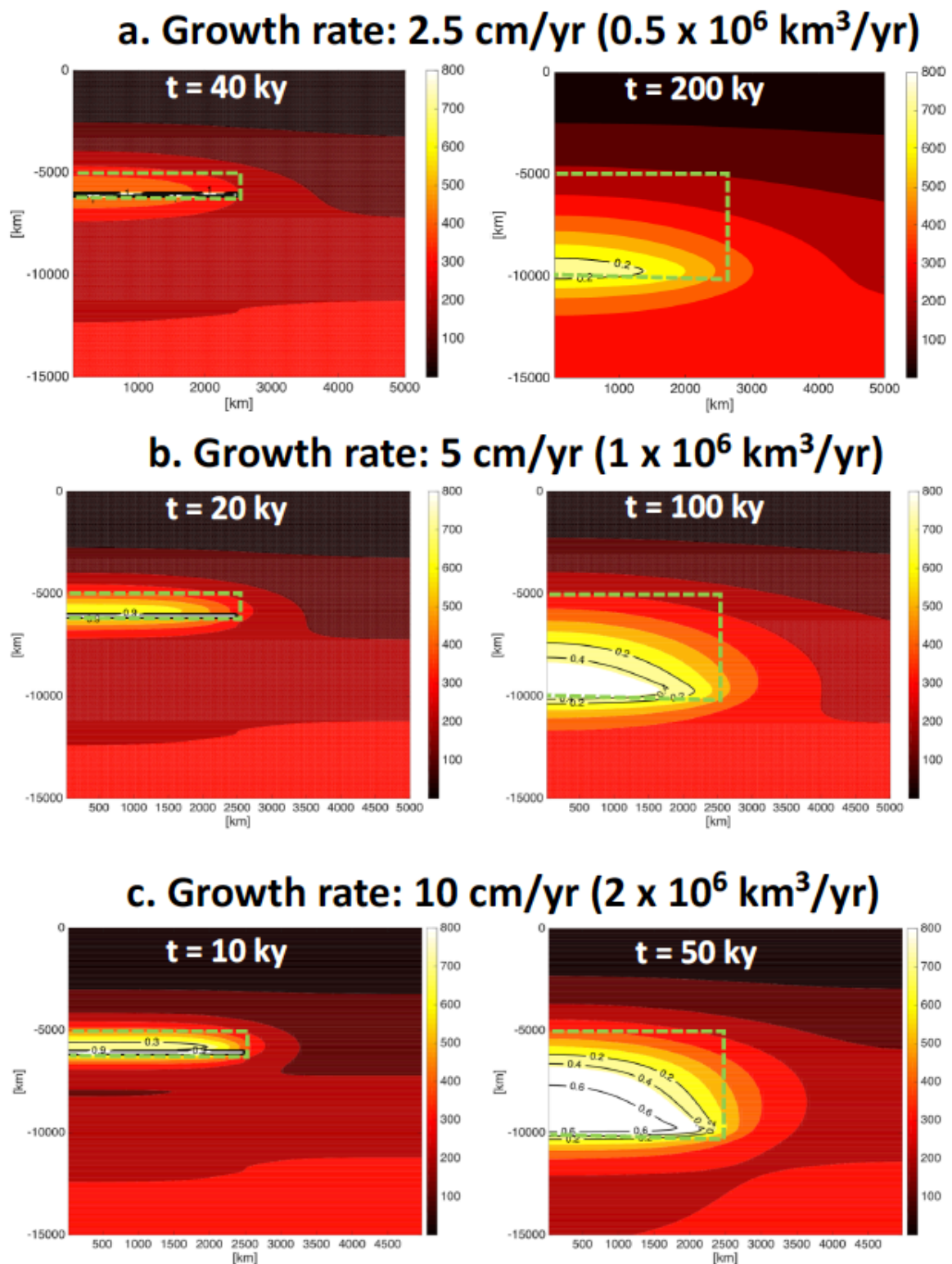


Figure 5.

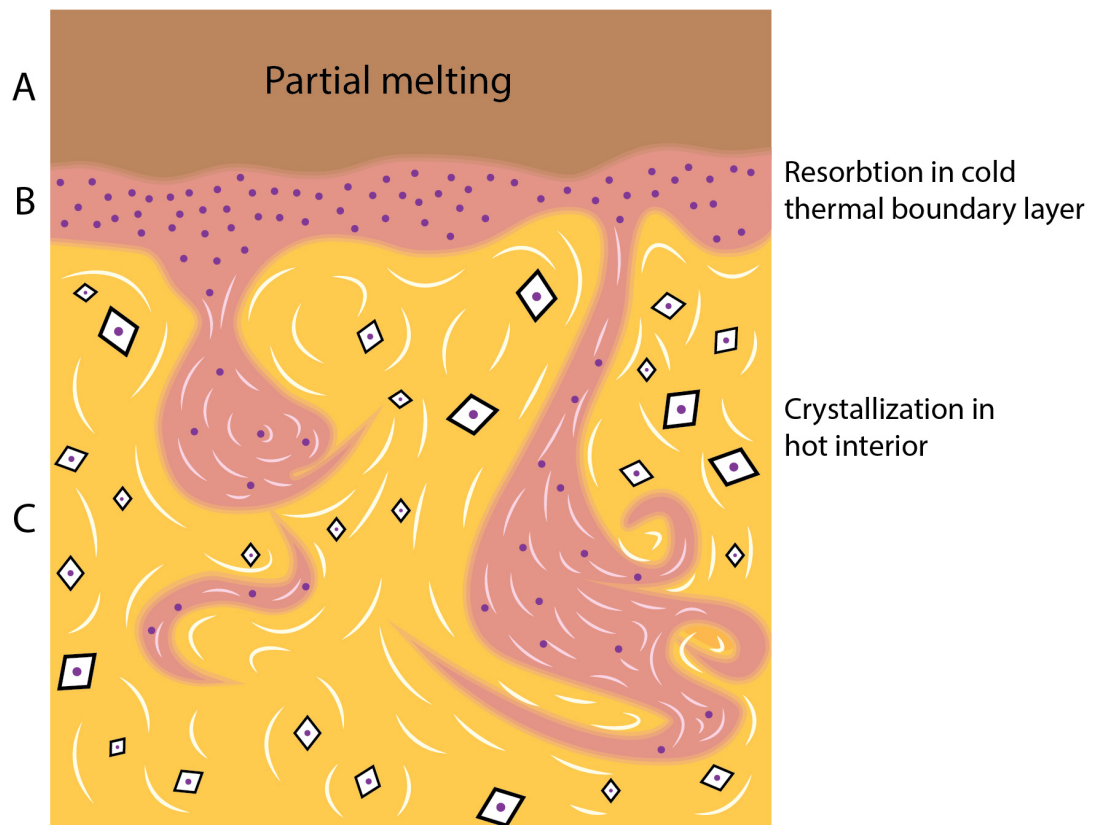


Figure 6.

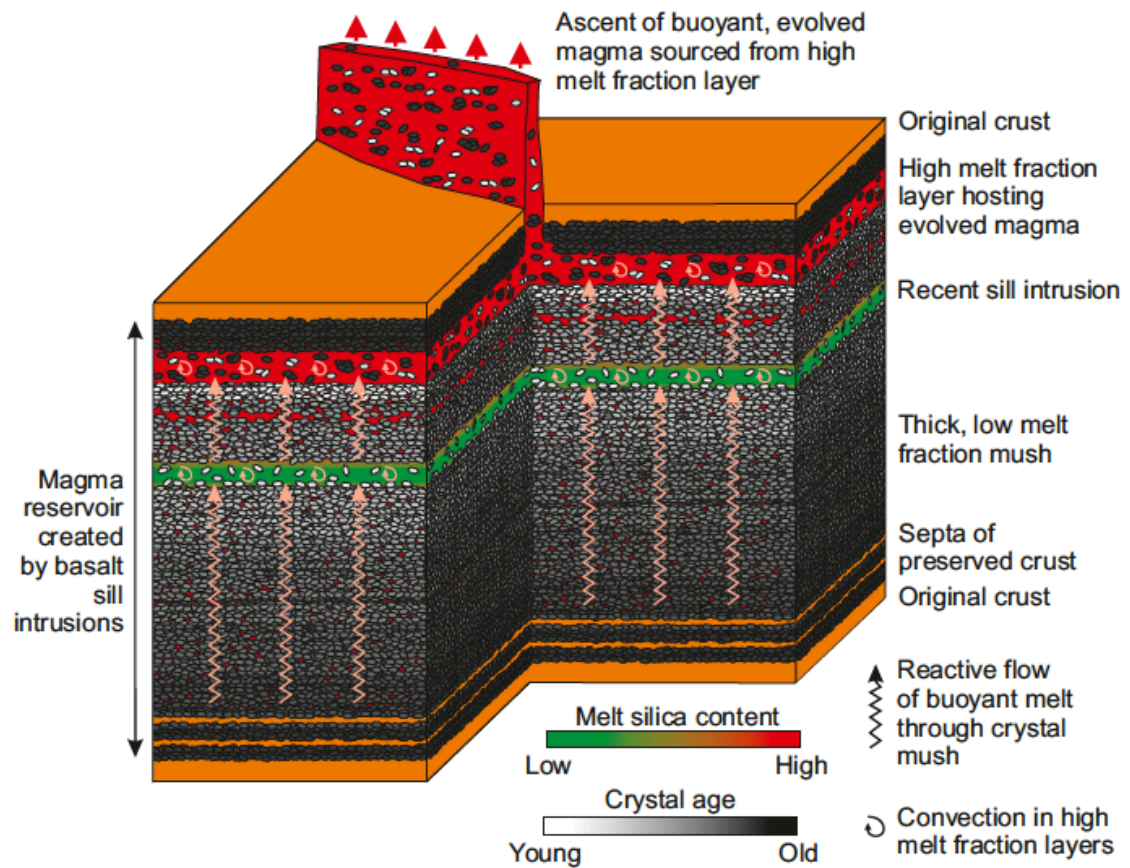


Figure 7

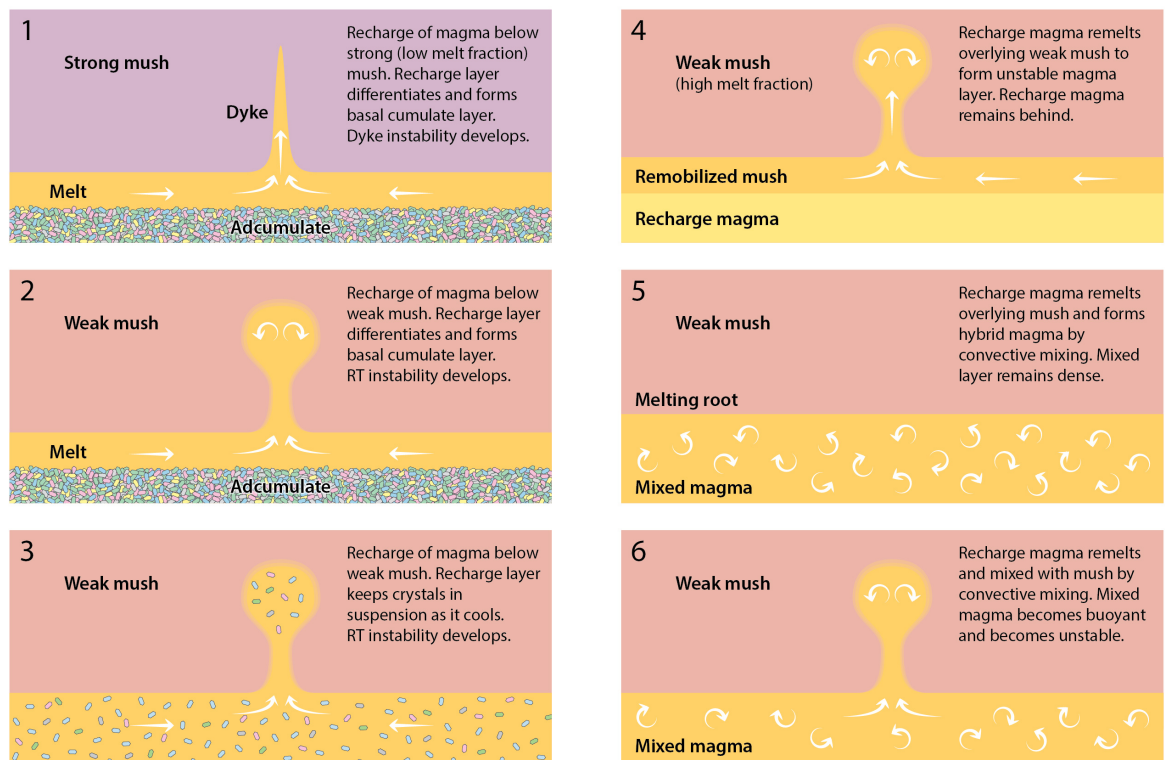


Figure 8.

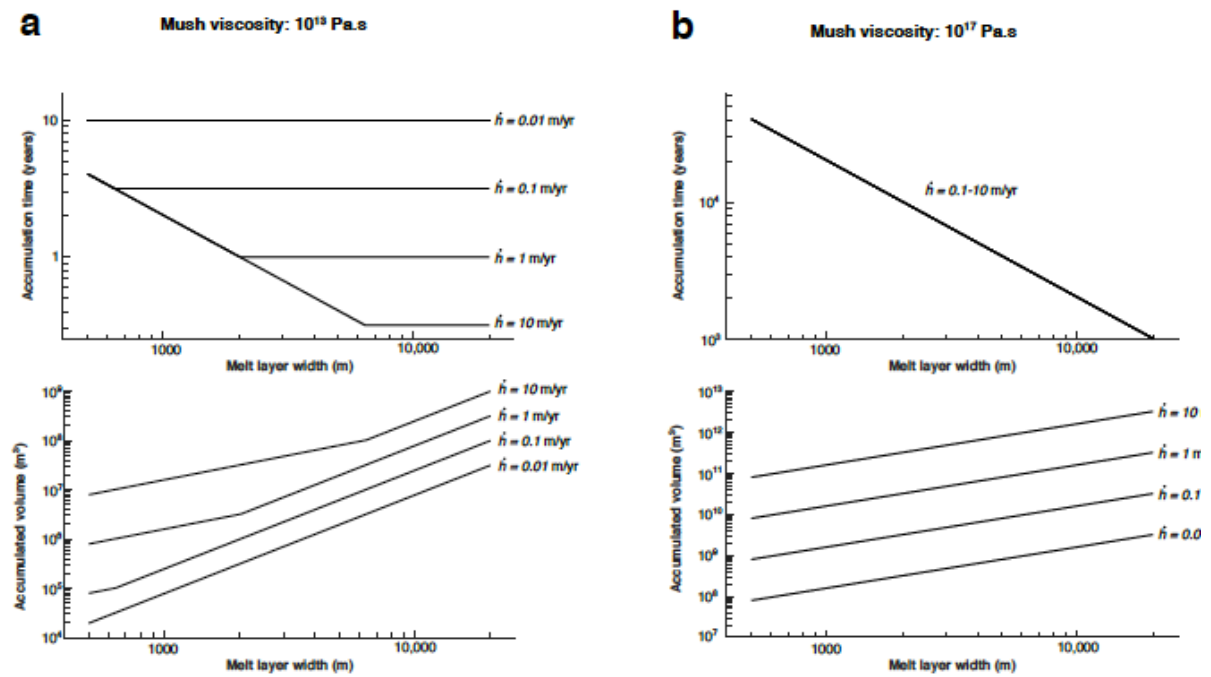


Figure 9.

